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Discussions

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# *Interactive comment on* "Quantifying the global atmospheric power budget" *by* Anastassia M. Makarieva et al.

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Dr. Garrett commented: "Latent heat release is considered the primary mechanism for cloud production, since by reducing density, it enables cloud parcels to be positively buoyant with respect to their surroundings. LES models of cloud development appear to reproduce cloud phenomena very well without accounting for any reduction in atmospheric volume due to condensation. What is missing?"

In brief, it depends on cloud type and spatial scale. Our view is that condensation as a circulation driver is least important for shallow convective clouds, comparable to buoyancy for deep convective clouds and dominates at horizontal scales exceeding the atmospheric scale height.

In parallel, Large-Eddy Simulations (LES) are more unambiguous and thoroughly tested by observations for shallow non-precipitating clouds (where net condensation is zero) than they are for deep convection. For precipitating clouds LES are less robust and require additional tuning. Finally, LES models, where the large-scale environmental properties are set up externally, cannot in principle quantify the large-scale dynamic effects of condensation, i.e. the upscaling of condensation effects remains unaccounted for.

While latent heat release does make cloud parcels positively buoyant, this does not only enhance the upward motion but also suppresses the subsiding motion. The latter is equally important for cloud formation. For example, if evaporative cooling in downdrafts (which compensates for the latent heat release in updrafts) is switched off, deep convection in LES models may not form at all. The entrainment of environmental air is crucial for clouds. The role of condensation dynamics in this process is implicit in the corresponding parameterizations.

Below we discuss these statements in greater detail.

## 1 Relevant scales for condensation dynamics

The basic spatial scale of condensation dynamics is the scale height $h_\gamma \equiv -\gamma/(\partial\gamma/\partial z)$ of the relative partial pressure $\gamma \equiv p_v/p$ of saturated water vapor. It is of the order of the atmospheric height $h_\gamma \sim h \sim 10$ km. This is why water vapor condensation can generate atmospheric motions[1].

If there were no pressure adjustment, condensation in the ascending air would create a pressure perturbation of the order of $\Delta p \sim p_v h_c/h_\gamma$, where $h_c$ is the vertical dimension

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[1] In contrast, evaporation cannot: here the relative partial pressure of water vapor varies over a microscopic scale of the order of one free path length above the evaporating surface (see Section 3.1 in our article). At this scale dominated by molecular viscosity macroscopic motions cannot arise by definition.

of the condensation region (Fig. 1a). If $h_c \ll h$, then the unperturbed pressure values $p_t$ and $p_s$ at the top and bottom of the condensation region are approximately equal, $p_t \approx p_s$. In this case the downward and upward pressure adjustment processes initiated by the pressure perturbation counteract each other and their cumulative impact on the ascending motion is negligible.

Likewise, if the horizontal scale $l_c$ of the condensation area is much smaller than the vertical one, $l_c \ll h_c$, the pressure adjustment processes will occur more rapidly in the horizontal plane than in the vertical plane. Since prior to adjustment the pressure perturbation would be maximum at the top of the condensation area, the horizontal air convergence at the cloud top could suppress the vertical motion and hence condensation itself.

Meanwhile if condensation occurs over a larger distance $l_c \gg h_\gamma$, the effect of the horizontal pressure adjustment is affecting only the edge of the condensation area and is thus minor relative to the upward vertical adjustment. Thus, condensation dynamics can act as a circulation driver at the scale $l_c \gg h_\gamma$ provided that $h_c \sim h_\gamma$ (Fig. 1b). This is the case of deep convection occupying a larger region with $l_c \gtrsim 100$ km.

## 2 Shallow convective clouds

In shallow convective clouds neither of the above two conditions is fulfilled. Instead, we have $h_\gamma \gg h_c \sim 2$ km and $h_c > l_c \sim 0.5$ km. The role of condensation as a possible driver of these clouds is minimal. Besides, in non-precipitating clouds condensation (in the cloud core) and evaporation (in the subsiding shell) compensate each other at a horizontal distance $l_c \ll h_\gamma$. In this case condensation dynamics does not have a direct impact on the larger scale circulation either. However, its within-clouds effects are implicitly included into the LES parameterizations of turbulence. We explain this below.

LES models have been primarily tested for shallow convective clouds (e.g., Rodts et al., 2003; Neggers et al., 2003; Jonker et al., 2008; Heus et al., 2009; Katzwinkel et al., 2014). A remarkable finding confirmed by observations was a subsiding shell surrounding each cloud. This finding altered the previous view of cumulus convection which presumed that subsidence occurred over a large area rather than was concentrated near the cloud (Jonker et al., 2008).

It was hypothesized that the subsiding shell can either be driven by negative buoyancy (i.e. when the subsiding air is colder than the ambient air) or by the mechanical forcing (i.e. by a relative pressure surplus at the cloud top that would be pushing the air down) (Rodts et al., 2003; Jonas, 1990).

In the latter case the subsiding shell could be positively buoyant – like the warm air descending in the hurricane eye. From the thermodynamic viewpoint (i.e. considering the circulation energy budget), such a motion is possible. If the potential energy associated with buoyancy (the conventional CAPE) is transformed into kinetic energy of the rising air and if this kinetic energy is then transformed into the potential energy of the pressure surplus at the cloud top, this pressure surplus could make the cloudy air descend even if it is relatively warm.

In reality, however, LES models and observations showed that the subsiding cell is driven by negative buoyancy caused by evaporative cooling of ambient air. There were no positively buoyant downdrafts. This means that the kinetic energy of the ascending motion dissipates to heat rather than transforms to the potential energy of the pressure perturbation. This property is set by the parameterization of the dissipative processes (i.e., turbulence) in the LES model. Thus *it is turbulence that determines to what degree the positive buoyancy of the ascending air is able to drive the circulation as a whole* (and to what degree this circulation depends on additional factors like air entrainment and evaporative cooling).

Evaporation can cool only those air parcels that have not been previously warmed by

latent heat release (if a droplet evaporates in the same air parcel where it condensed the net effect will be zero). Thus, given the dominant role of negative buoyancy, the entrainment of external air into the cloud is crucial for cloud formation (de Rooy et al., 2013). Condensation initiates pressure adjustment processes and thus impacts the entrainment process both in the vertical and horizontal plane (Fig. 1a). This impact should be implicitly taken into account into those parameters of turbulence that are fitted to observations.

In summary, large-scale air flow interacting with the planetary boundary would produce shallow convection in both dry and moist atmosphere. In this sense condensation does not directly drive shallow convection. However, the dynamic effects of condensation modify the properties of the shallow clouds by impacting entrainment and turbulence.

## 3  Deep convection

For deep convection we have $h_\gamma \sim h_c$ and $l_c \lesssim h_c$ for individual clouds. It is therefore the minimal horizontal scale where condensation can act as the driver of circulation. LES studies of deep convection demonstrate that by choosing a proper turbulence scheme, spatial resolution and cloud microphysics it is possible to quantitatively describe some cloud properties (Khairoutdinov and Randall, 2006; Morrison et al., 2015; Heath, 2015; Potvin et al., 2017; Fiori et al., 2017). However, what determines the correct choice of key parameters remains uncertain.

Thus, current LES models of deep convection do not prove that condensation dynamics is unimportant compared to buoyancy. With maximum air velocity of the order of $\sqrt{2p_v/\rho} \sim 60$ m s$^{-1}$ available from condensation, to drive convective clouds with their typical velocities of 1-2 m s$^{-1}$ requires only a minor portion of total potential energy associated with the partial pressure of water vapor. Compared to observations, deep convection simulated without accounting for condensation dynamics might, for

example, display an excessive positive (negative) temperature anomaly of the updrafts (downdrafts). Furthermore, condensation dynamics, with its non-equilibrium vertical pressure gradients, can play a significant role in overcoming convective inhibition. This cannot be elucidated by current LES models.

Most importantly, in deep convective clouds LES outputs are crucially dependent on the parameters of the large-scale circulation that are externally imposed onto the LES model. (As a simple example, a large-scale subsiding motion will suppress deep convection.) Without an explicit account of condensation dynamics it is not possible to correctly quantify the feedback of deep convection on the large-scale motion and hence to obtain a self-consistent picture of atmospheric circulation.

## 4   What is missing on a larger scale?

Potential energy available for conversion to kinetic energy is associated with spatial heterogeneity. In the conventional approach outlined by Lorenz, available potential energy arises when some parts of the atmosphere are warmed (or cooled) more than the others. If the atmosphere is uniformly warmed or cooled, potential energy is not available.

The situation is similar for condensation. Consider a horizontally isothermal atmosphere in hydrostatic equilibrium. We remove some gas uniformly across the entire atmosphere in such a manner that the hydrostatic equilibrium is not perturbed (i.e. we remove a constant air fraction at each altitude.) In such a case pressure declines everywhere, but no motion results.

Now consider removing gas by condensation and precipitation from a large but limited region of horizontal size $l_c \gg h$ (e.g. the equatorial region), once again without perturbing the hydrostatic equilibrium. Now, as the region's surface pressure declines, air will flow towards it from the surrounding atmosphere. The greater the surface pressure

perturbation $\Delta p_s$, the greater the cross-isobaric horizontal velocity $u$ of the air flow.

Historically though meteorological sciences attributed observed $\Delta p_s$ to differential heating assuming its dominant role in driving atmospheric circulation. Since it was not possible to quantify $u$ and $\Delta p_s$ from theory knowing $\Delta p$ in the upper atmosphere (Fig. 2a,b), the values of $u$ and $\Delta p_s$ were fitted to observations by adopting the necessary parameterizations of turbulence (see Introduction in Makarieva et al. (2017) for a more detailed discussion).

Having thus built a plausible model of dry atmospheric dynamics, people then added a "mass sink" via the moist continuity equations, which specify how precipitation influences the surface pressure tendency in a hydrostatic atmosphere. Since little changed in the resulting circulation patterns, it was concluded that the "mass sink" (and, hence, condensation changing the amount of gas) is inconsequential.

However, this overlooked the main effect of condensation: the formation of a local non-equilibrium vertical pressure gradient in air parcels rising in the gravitational field. As we discussed in our previous comment[2], it may have been overlooked because this gradient elicits a pressure adjustment and vanishes on time scales significantly shorter than the characteristic time scales of the large-scale air circulation it generates. Indeed, condensation is not equivalent to a hydrostatic mass sink. Any new droplet instantaneously produces an upward pressure gradient: since oversaturation of water vapor in a dry adiabatically ascending air parcel increases with height, condensation removes more gas from the upper part of any volume affected by droplet formation than it does from the lower. If $h_c \sim h_\gamma$ the process of hydrostatic adjustment affects the entire atmospheric column (Fig. 1b), enhances the ascending motion and transforms potential energy contained in the condensation-induced pressure perturbation into the kinetic energy of macroscopic air motions.

Assuming that it is condensation that drives the circulation explains, first, why the char-

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[2] https://doi.org/10.5194/acp-2017-17-AC3

acteristic horizontal pressure perturbations $\Delta p_s$ coincide in the order of magnitude (10 hPa) with the partial pressure of water vapor at the planetary surface – and are independent of the horizontal scale of the circulation. Indeed, in the words of Holton (2004), apart from the synoptic scale of $10^3$ km "pressure fluctuations of similar magnitudes occur in other motion systems of vastly different scale such as tornadoes, squall lines, and hurricanes". Second, condensation dynamics also explains the observed relationship between the circulation *power* and precipitation in phenomena as diverse as hurricanes (e.g., Makarieva and Gorshkov, 2011) and global atmospheric circulation (Makarieva et al., 2013b,c, and our present article).

Put simply, in the conventional picture the temperature-induced pressure gradient pushes the air away from the warmer air column in the upper atmosphere (Fig. 2b). This creates a relative pressure deficit in the upper atmosphere and initiates the upward motion. Our position, on the other hand, is that the pressure gradient in the upper atmosphere cannot ensure the necessary pressure deficit in the vertical (because in a rotating atmosphere a steady state is the geostrophic air flow with no cross-isobaric motion). It is a dynamic (not a thermodynamic) limitation[3]. The ultimate cause of the circulation is the ascent induced by the condensation pressure perturbations (Fig. 2c).

Since condensation usually occurs in warm rising air (although there are exceptions like the Ferrel cell), turbulence parameterizations fitted to support the temperature-driven model often produce a realistic output. However, if, as we argue, the real driver of the circulation is not temperature but condensation, then such models will fail to predict what happens when the considered area gets warmer but drier (i.e. condensation and warmth do not coincide in space and time). This situation is especially relevant for the prediction of monsoons and the effects of deforestation. If one assumes that it is temperature that drives winds (while it is not), we underestimate the danger of deforestation for the atmospheric transport of moisture from ocean inland (Makarieva

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[3]Another dynamic limitation, as we discussed in the previous section, does not allow deep convection to form unless there are negatively buoyant descending air parcels.

and Gorshkov, 2007; Makarieva et al., 2013a).

To summarize, condensation dynamics appears to have been neglected without a serious assessment of its crucial aspects. We believe that without accounting for these aspects the challenges currently faced by the meteorological science when describing moist atmospheric dynamics will persist. We thus welcome this discussion, thank Dr. Garrett once again for having made is possible and look forward to its continuation in the future, in one form or another.

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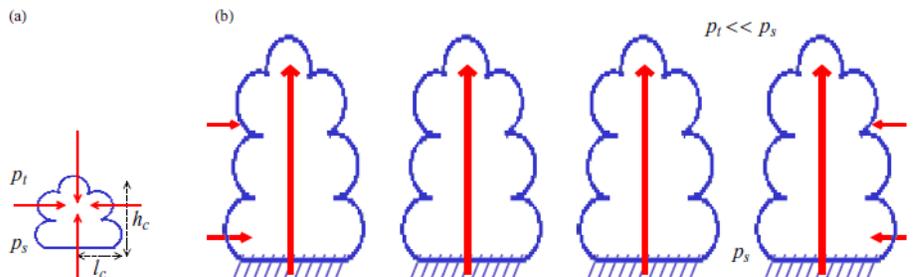
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Discussion paper

Figure 1: Scales relevant for condensation dynamics, see text for details. (a) Shallow non-precipitating cloud, (b) a large area occupied by deep convection. Red solid arrows indicate the direction of air motion during the pressure adjustment induced by condensation.
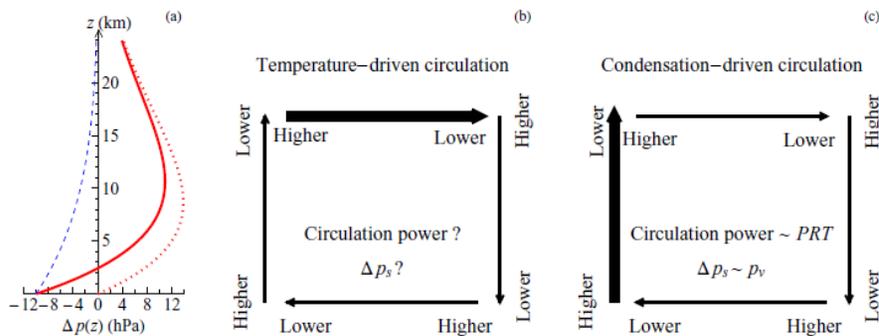
**Fig. 1.**

Figure 2: Different views on what drives the large-scale atmospheric circulation. (a) Vertical profile of the pressure difference between two hydrostatic air columns of which one is warmer by 10 deg K but has a pressure deficit of 10 hPa at the surface (thick solid curve), see Fig. 1d of Makarieva et al. (2017). (b) Temperature-driven circulation: the necessary condition for the circulation to occur is the temperature-induced horizontal pressure gradient in the upper atmosphere (thick horizontal arrow), see panel (a). (c) Condensation-driven circulation: the necessary condition is the vertical pressure perturbation associated with condensation in the ascending air (thick vertical arrow). "Higher" and "Lower" at horizontal arrows refer to air pressure relative to the average at that height (cf. Fig. 2.6 of Vallis (2006)); "Higher" and "Lower" at vertical arrows refer to the corresponding perturbations of the hydrostatic pressure distribution.

**Fig. 2.**